Quantifying, predicting and verifying changes in soil carbon

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Introduction

Soils contain more than twice as much carbon as the atmosphere on a global basis. Thus the exchange of carbon between land and the atmosphere has a critical influence on the amounts of CO_2 in the atmosphere and on organic matter in surface soils. Concern about climate change has led to considerable interest in the potential for mitigating increasing atmospheric CO_2 by sequestering organic carbon in soils. Soil scientists have long been concerned about changes in soil organic matter associated with changes in land use (e.g. perennial native vegetation converted to annual cropland) or agronomic management (e.g. contrasting cropping systems). Despite the history of research on soil organic matter, assessments of potential carbon sequestration still are constrained by the availability of suitable methods to quantify, predict and verify changes in soil carbon. This paper summarizes the strengths and weaknesses of the best available methods and tools.

Measuring soil carbon

Sampling soil and calculating carbon storage

Soil C storage depends on soil area and thickness or mass. Areal inventories of vegetative cover (or non-vegetation, such as ice or rock), land use (forest, agriculture, etc.), and land management (e.g. plantation forest, annual cropping, perennial forage, native rangeland) are required to estimate soil C on a regional or national basis. The starting point for such estimates is measurement of soil C stored per unit area at specific points in the landscape. At these points, soil samples of a known volume are collected and the organic C concentration is analyzed. The mass of soil organic C per unit area for a specified soil thickness may then be calculated by multiplying the soil organic C concentration by soil bulk density:

$$\underset{kg\ C\ m^{-2}}{\text{Soil C,}} = \underset{kg\ C\ Mg^{-1}\ soil}{\text{C concentration,}} \times \underset{Mg\ soil\ m^{-3}}{\text{bulk density,}} \times \underset{m}{\text{thickness,}} \quad [1]$$

Clearly, the mass of soil C per unit area is dependent on the thickness of soil under consideration, thus any estimate or inventory of soil carbon must explicitly indicate the depth of soil under consideration.

The concentration of organic C in most soils generally decreases with increasing soil depth, because inputs of plant residues, which are responsible for the accumulation of soil organic matter, usually are localized at the soil surface. Changes in land use (e.g. native vegetation versus annually cultivated cropland) and agronomic management (e.g. conventional versus conservation tillage) typically have a greater influence on soil carbon in surface layers than sub-surface layers. Consequently, the change in soil C, expressed as a percentage of the quantity present initially, tends to decrease as the depth of soil under consideration increases. The soil depth for evaluating carbon change traditionally has been left to the discretion of the researcher who might base the decision on factors such as cultivation depth, study objectives and duration, depth of root penetration and solum thickness. To account for changes in C concentrations with soil depth, the total sampling thickness usually is sub-divided into several layers, and the total soil C mass per unit area is calculated as the sum of C stored in successive soil layers to a specified depth.

For a given C concentration, the quantity of soil C per unit area also depends on soil bulk density. To calculate the bulk density of gravely or stony soils, the mass of coarse fragments contained in the volumetric sample is subtracted from total sample mass, because coarse fragments (>2 mm) are excluded from most soil chemical analyses. Together the thickness and bulk density of a soil layer determine the mass of soil under consideration, and soil C mass per unit area is calculated as the product of soil mass (determined by the sample dimensions and dry weight) and C concentration (analyzed in the laboratory). Investigations of management-induced changes in soil C storage traditionally have neglected the influence of soil mass on comparisons among contrasting soils. Most investigators erroneously compare C stored in layers of equal soil thickness, whereas accurate comparisons should be among layers of equivalent soil mass. Management practices, such as tillage and packing, often have a considerable influence on soil bulk density. Consequently, for the same analyzed C concentration, soils with the

greater bulk densities contain greater masses of soil and C than equal thickness' of soils with lesser bulk densities. The mass of soil C in an equivalent or standard soil mass, per unit area, may be calculated by estimating the thickness of the deepest soil layer required to attain the equivalent mass, and summing the mass of soil C contained in this and the above layers (Ellert and Bettany, 1995).

Forms of soil C and their analysis

A distinction between plant residue C and soil C must be established at the outset of any attempt to measure changes in soil C. Soil C traditionally has been regarded as that associated with dry soil which has been crushed to pass a sieve with 2 mm openings. Methods to dry and crush the soil rarely are standardized and exclude a variable amount of plant roots and soil-incorporated stems which often are discarded as non-soil. Clearly this is an arbitrary definition of soil which does not facilitate quantitative assessments of C storage under rival management practices, often differing only in the extent of residue incorporation into the soil. A more meaningful approach to assess total ecosystem C pools is to distinguish between above-ground C (possibly sub-divided, as the case requires, into standing live phytomass, anchored dead litter, and loose surface residues) and soil C. This approach requires collection of the above-ground C first, and whatever remains on and in the soil is included as soil C. Since soil C thus defined also includes plant crowns in the surface layer, the ratio of plant crowns to bare soil in the area included in the soil sample should represent that of the ecosystem. The soil C contained in such samples may be subdivided using various methods to define physical fractions or functional pools of soil or below-ground C which provide insight to C cycling processes.

Several fractions of soil organic C have been defined on the basis of chemical, biological and physical characteristics. Chemical fractions differentiate according to the chemical extractability which is thought to reflect the structural chemistry of the organic compounds. Biological fractions are differentiated according to bioassays of decomposability or identification of organic C within microbial biomass. Physical fractions are differentiated according to the size, density or location (e.g. aggregated with mineral particles or coating soil peds or faunal burrows) of organic particles. Fractionation methods tend to be well documented, but results may be difficult to

interpret if they pertain to an arbitrary portion of below-ground C and information on initial collection and handling of the soil samples is neglected. Certain 'actively cycling' fractions of soil organic C are more sensitive than total organic C to the influence of management, because the quality or susceptibility to biological transformations is distinct from that of the remaining soil organic C. In such cases, the proportions of soil organic C within actively cycling fractions should be calculated to distinguish the effects of management on organic C quality from those on total organic C quantity.

The fundamental distinction between organic C and inorganic C is an important one for most soils, and easily made by chemical analysis in the laboratory. Automated instruments are commercially available for estimating total soil C by dry combustion, and some instruments simultaneously determine total N using the Dumas method. Older methods based on wet oxidation to determine organic matter are not recommended as they are less reliable, more time consuming, and generate more toxic waste. Since soil organic matter is poorly defined and has a variable C content, analysis and reporting of soil organic C is more accurate and meaningful, as it quantifies the C derived from organic compounds. Soil organic C (SOC) may be analyzed directly by combusting samples which have been acidified to eliminate inorganic C or by combusting samples under conditions to selectively oxidize organic C to CO_2 without decomposing inorganic C. SOC may also be determined as the difference between separate analyses of total and inorganic C. Both of these approaches include charcoal, coal-like materials and elemental C as SOC, because such materials are recovered by high temperature combustion and persist after acidification.

Most analytical methods for soil inorganic C involve acidification of the sample and analysis of liberated CO_2 by titration, gas chromatography, manometry, or weighing. Weathering or formation of soil inorganic C, which consists of carbonate minerals, may generate or sequester CO_2, but in most soils these processes have a minor influence on CO_2 exchange relative to that of biological processes involving soil organic C (Curtin et al., 1994). Thus the primary goal of soil inorganic C analysis is to estimate SOC by subtraction from total C. Accounting for the presence of soil inorganic C is especially important in sub-surface layers where carbonate from the upper layers accumulates during

pedogenesis. Analysis of both soil C and N is useful to confirm SOC measurements, because the ratio of SOC to total N rarely exceeds 10 in lower soil layers, with greater ratios often indicating an incomplete distinction between organic and inorganic C.

Soil carbon variability

Spatial variations

The spatial variability of soils is consistent with the complexity of nature. An understanding of SOC variability is imperative to estimate C storage in complex landscapes and to distinguish spatial from temporal changes in C storage. Systems of soil taxonomy recognize spatial variations associated with differences in factors such as climate, plant community, topography, parent materials and human activity. Land resource databases contain valuable information on variations in SOC concentrations among broad taxonomic groups distributed over hundreds of km (e.g. Tarnocai and Lacelle, 1996). Soil variations, including SOC contents, along catenas or toposequences have been recognized and used to study soil processes under contrasting environmental conditions. More recently researchers have been quantifying soil spatial variability at scales of less than one km by collecting samples at grid points superimposed on complex landscapes (e.g. Pennock et al. 1994). Such studies have built on qualitative data from earlier work on toposequences, and improved understanding to the extent that generalizations may be made about the distribution of soil C storage within variable landscapes.

Soil redistribution

The mass of organic C present in soil is a function of biological processes and geomorphological processes. The biological processes of photosynthesis and respiration determine the balance between plant C inputs to soil and CO_2 outputs from decomposition of SOC. The geomorphological processes of soil redistribution by wind, water, tillage and gravity influence SOC mass by removing soil and associated C from one area and depositing it on another area. Such processes usually increase the spatial variability of SOC. For example, in Chernozemic soils on hummocky terrain, plant C inputs and thus SOC generally are lower on the knolls, because water runs to lower slope positions. When such soils are cultivated to produce annual crops, erosion by tillage, water and wind removes the already thin topsoil from the knolls and deposits it at lower slope positions (e.g. Martz and deJong, 1987). This zero net erosion or loss of soil from the landscape (erosion from knolls ~ deposition at lower slope positions) is typical of the Canadian Prairies where a major proportion of the area is internally drained (Lemmen et al., 1997) and drainage basins are closed. In regions such as the St. Lawrence lowlands, drainage basins tend to be more open, so that there may be positive net erosion or export of soils from terrestrial systems to rivers, lakes and ultimately the sea.

Regardless of whether or not there is net erosion from a landscape, the influence of soil redistribution (within landscapes or from land to sea) on the conversion of SOC to atmospheric CO_2 is uncertain. From a management perspective, the negative impact of soil redistribution on productivity and farm operations usually justifies prevention measures, regardless of the consequences for atmospheric CO_2. Whether measures to prevent soil erosion or redistribution reduce or enhance SOC sequestration is difficult to ascertain. Estimates of management-induced changes in SOC storage at specific sites in the landscape may be simplified by selecting sites with negligible soil redistribution, otherwise careful measurement of redistribution is required to account for differences in masses of soil and associated C. If temporal differences in soil mass at eroded sites are ignored, actual SOC losses will be partially compensated and thus obscured by inclusion of deeper layers. Conversely, at sites with deposition, SOC gains will be underestimated and thus obscured by omission of deeper layers.

Measuring temporal changes

Accurate assessments of temporal changes in SOC storage provide valuable information on the net exchange of C between land and the atmosphere. Such changes reflect the combined influence of several sporadic and fluctuating fluxes influenced by complex interactions among several processes and environmental variables. Although less demanding than calculating C budgets from detailed flux estimates, several factor make it challenging to measure temporal changes in SOC storage. The annual imbalance between soil C inputs and outputs (rarely exceeding 5 Mg C ha^{-1} yr^{-1}) is small relative to SOC storage (usually more than 50 Mg C ha^{-1} in the 0-30 cm layer). Changes in SOC which might be anticipated for a modification of agronomic management on existing cropland are smaller than historical changes which occurred when native forest or grassland was cleared and broken for agricultural use. As a consequence of the large area of Canadian cropland (on the order of 45 million ha), even a very small change in SOC (e.g. 0.2 Mg C ha^{-1} or 9 million Mg C) may amount to an important proportion of the reduction target for CO_2 emissions (8 million Mg C ~ 6% of Canada's CO_2 emissions in 1990, although this has increased since). An obvious strategy to assess temporal change associated with management is to quantify SOC at sampling times separated by three or more years. This allows the effects of annual imbalances between soil C inputs and outputs to accumulate and smoothes over seasonal fluctuations.

Estimates of management-induced changes in SOC storage may be influenced by both spatial and temporal variability. To ensure that changes in SOC represent a temporal rather than a spatial shift, the sampling points for the initial measurement may be interspersed with those for the subsequent measurement, such that spatial differences between sampling times are comparable to those within sampling times. The temporal change may then be estimated as the difference between the initial and subsequent measurement. Since this temporal change includes the influence of both management and temporal variations (e.g. those associated with weather), carefully designed experiments include check or control treatments as well as one or more treatment with modifications or improvements to management. The influence of management may be determined unambiguously by comparing SOC stored in samples collected at the same time from

contrasting management treatments. In regional studies to monitor management-induced changes in SOC it may be unfeasible to include control treatments, but it must be recognized that any temporal changes may include variability which is independent of management. Since independent variability may accentuate or diminish the apparent influence of management, monitoring studies ideally should be designed so that management effects are the dominant contributor to temporal change.

Isotopic assessments of soil C changes

Changes in SOC are determined by the balance between production and retention of plant residues, but quantitative data are scarce. Measuring the amount of plant residues produced and retained in any given year is difficult, because fresh residues become mixed and diluted within a much larger (typically more than ten fold) pool of SOC, some of which had been deposited thousands of years earlier. Estimating below-ground production of plant C is complicated by errors in measuring roots and exudates and their rapid decomposition. Plant C allocated below-ground often is estimated from that measured above-ground, but the relationship is highly variable and is influenced by several factors including: plant species, growth stage, soil conditions (fertility, texture, structure), and climate. Hamblin et al. (1990) reported that the percentage of spring wheat dry matter allocated below ground by maturity ranged from 10 to 50%, depending on N availability and moisture stress. Over a period of several years, however, the net amount of plant C retained over a period of several years may be more important than actual plant C inputs, below- and above-ground, in any particular year.

Differences between the isotopic composition of plant C and SOC allow distinction between the two pools and estimation of plant C inputs and retention in the soil. Isotopes are different forms of an element with very closely related chemical properties and the same atomic number but different mass numbers (i.e. the same number of protons but different number of neutrons). The most common isotopes used in carbon cycling research are ^{12}C, ^{13}C and ^{14}C. The two most common C isotopes in nature are ^{12}C (typically 98.89 %), and ^{13}C (about 1.11%), both of which are stable. ^{14}C or radiocarbon accounts for only a small fraction of natural C (on the order of 1×10^{-10} %), and is

unstable, meaning that it decays (to ^{14}N with a half-life of 5730 years) with the release of radiation. In many ecosystems the isotopic composition of SOC equilibrates with that of the plant C from which it is derived, and thus the two pools are isotopically indistinguishable. In other ecosystems or controlled environments plant C is isotopically distinct from SOC, and isotopic variations are a powerful tool to elucidate the processes contributing to SOC change.

Intentional manipulation of isotopic composition

Perhaps the most widely used approach to study the fate or retention of C in any organic compound added to soil is to artificially synthesize the compound with an unnaturally high enrichment of ^{14}C. When such compounds are added to soils with natural levels of ^{14}C, the extent to which C from the compound is retained in the soil or released as CO_2 may be determined by analyzing the ^{14}C content of the soil and evolved CO_2. More elaborate systems have been devised to feed growing plants with ^{14}C-labeled CO_2 so that the plant C, including roots and exudates, is isotopically distinct from the SOC. Such systems tend to be technically demanding to maintain and operate, with common problems including: maintenance of air temperature and humidity at realistic levels to support plant growth within a closed space with the isotopic composition of CO_2 at artificially enriched levels, uniformity of the isotopic composition of various plant parts throughout all growth stages, maintenance of a rooting environment (temperature, structure, moisture distribution, depth) which represents natural growing conditions, and rapid C transformations which tend to obscure net annual dynamics of plant C (e.g. decay of fine roots producing ^{14}C-labeled CO_2 and microbial biomass in the soil, plant uptake of unlabelled CO_2 from SOC decay). Despite these and other limitations, the estimated proportions of plant C allocated below-ground in enriched isotope tracing studies still form the basis for estimating below-ground and total net primary productivity from measurements of above-ground production in many ecosystems.

Recent advances in instrumentation now enable the use of CO_2 enriched in ^{13}C rather than ^{14}C. The considerable gains in safety from using a stable isotope without radiation hazards (both personal and environmental) are somewhat offset by the greater amounts of plant ^{13}C required to produce detectable changes in soil ^{13}C, and by the higher

analytical complexity and costs (mass spectrometry instead of liquid scintillation counting of decay events). The use of isotopically distinct CO_2 in free-air CO_2 enrichment (FACE) studies has allowed the tracing of plant C (slightly depleted in ^{14}C and often ^{13}C relative to plants grown in the normal atmosphere) to soils in field plots without the effects of the artificial environments typical of closed or partially closed chamber studies. Since the isotopic composition of plant C is only slightly different from that of SOC, isotopic assessments in FACE studies provide more reliable information on the retention of plant C over several years, rather than on the quantity of plant C produced in any particular year.

Natural and unintentional variations in isotopic composition

The isotopic composition of plant C may also be distinct from that of SOC as a result of natural or unintentional variations in ^{14}C or ^{13}C that allow tracing of C dynamics without artificial modifications. Traditionally, natural variations in isotopic abundance have been used to study geochemical processes, such as sedimentation and prehistoric environmental changes. Only recently have natural variations in isotopic abundance been used to study more contemporary biological processes in agricultural and native ecosystems. Since natural variations tend to be much less than intentional variations from artificial enrichment, natural variations usually are studied to evaluate the cumulative influence of several years of plant growth on SOC dynamics.

Natural variations in the ^{14}C content of SOC originally were used to assess soil age by radiocarbon dating in studies of pedogenesis. Later it was recognized that for most soils radiocarbon dating was a misnomer, because most soils actively exchange C with other biogeochemical pools (thus negating a fundamental requirement for radiocarbon dating), including the atmosphere, hydrosphere, lithosphere and biota. This did not detract from the value of using natural variations in soil ^{14}C, but contributed immensely to an understanding of differences among the decomposability of contrasting fractions of SOC. Studies of natural variations in soil ^{14}C generally show the extent to which recently added plant C decomposes faster than the bulk of the SOC which has accumulated over the millennia. The unintentional pulse of ^{14}C released to the atmosphere by above-ground tests of nuclear weapons during the 1960s was assimilated

as CO_2 during photosynthesis and then deposited in the soil. As noted by Jenkinson, et al. (1992) this provided "an opportunity, hopefully unrepeatable, to follow the turnover of [soil] organic matter under natural conditions."

Natural variations in the ^{13}C content of plant C may reflect differences in atmospheric CO_2 (e.g. both concentration, and subtle increases in ^{13}C associated with fossil fuel combustion), growing conditions (e.g. water supply, position in vegetative canopy), and plant biochemistry. The most widely used natural difference in plant ^{13}C used to trace SOC change is that between plants using the C3 (CO_2 uptake catalyzed by Rubisco) and C4 (CO_2 uptake catalyzed by phosphenol pyruvate carboxylase) photosynthetic pathways. Since C3 plants discriminate more strongly against ^{13}C in atmospheric CO_2 than do C4 plants, the C3 plants are consistently more depleted in ^{13}C. Consequently, plant residues are isotopically distinct from SOC wherever the plant community is changed to one with a photosynthetic pathway which differs from the one where the soil originally developed. For example, the retention of plant C from *Zea mays* (a C4 crop) may be determined for soils which historically had developed under C3 vegetation and more recently were cropped to corn.

Natural shifts in the isotopic abundance of SOC are based on comparisons among samples collected from individual soils at successive time intervals or more often from adjacent soils (e.g. under C3 and under C4 vegetation) at a single time interval. Collections of stored soil samples obtained periodically from well documented and continuously maintained field experiments are a valuable tool for elucidating SOC change a temporal scales of 10 years or more. Since the pulse of bomb-^{14}C has influenced most soils collected after 1960, sample series which include both pre- and post-bomb soils are useful to elucidate the fate of bomb-derived plant C. Isotopic studies based on comparisons between adjacent soils collected at one time usually treats one site as the baseline or reference condition while the others (e.g. sites with vegetative cover changed to or from plants using the C4 pathway) have received inputs of plant C which is isotopically distinct from SOC. In these studies it is often assumed that the reference site represents the isotopic composition and sometimes quantity of SOC at the other sites before the isotopic shift in plant C occurred. This assumption should be valid under

steady state conditions (i.e. all sites were comparable before the isotopic shift in plant C, and there has been no change in SOC of the reference site since then), but may be dubious for some comparisons.

Modeling soil carbon

Mathematical models are imperative to clearly represent concepts about carbon cycling and to predict changes in soil carbon storage. Most scientists use conceptual models to assemble and communicate an understanding of nature, and mathematical models merely restate or extend this understanding. Despite this, some experimentalists and non-theorists often are suspicious of mathematical models which might confuse 'computer code and equations' for the 'real thing'. The cliché often applies that "no one believes the model except the person (theorist or modeler) who built it, and everyone believes new data except the person (experimentalist) who obtained it". Rather than representing hard facts, data are refracted through equipment and current theory, and scientific understanding of the 'real thing' (e.g. carbon cycling within agroecosystems) remains far from complete. Mathematical models are attempts to formalize and combine concepts about systems so that system behaviour might be reproduced. Any attempt to estimate changes in soil C storage on a regional basis must involve the use of mathematical models, because the spatial complexity of most ecodistricts render it impractical to measure soil C directly.

Scope and complexity of mathematical models

Mathematical equations to represent soil C changes have been used since the early 1900s, and likely before then (e.g. Salter and Green, 1933; Woodruff, 1949). The use of digital computers to perform dynamic simulations, typically involving several interacting rate equations which are integrated numerically is a comparatively young science (e.g. Garfinkel, 1962) which soon was applied to model soil C (e.g. Russell, 1964). Dynamic simulation models allow for interactions among several processes within the system. For example, soil organic C is dependent on plant production, which in turn depends on plant-available soil N, and all three of these are influenced by environmental variables, such as temperature and rainfall. Most complex models include some of the non-linearity

inherent in real systems (e.g. the circuitous cause and effect relationship between on plant-available soil N and plant growth). Ecosystem models have been classified according to various criteria, including: stochastic vs. deterministic; mechanistic vs. empirical or functional or correlative; simulation or analytic; hierarchical levels considered (Loomis et al., 1979; Baier, 1979; Addiscott, 1993). Most soil organic C models are deterministic, meaning that a certain set of conditions generates a unique outcome, whereas crop models used for planning and management often include weather as stochastic variables.

The scope and complexity of models varies with the number of dimensions, components, processes, and hierarchical levels considered. The goal of most soil C models is to describe the temporal dimension, but many models consider interactions among individual soil layers (vertical dimension in space). Few soil C models explicitly consider lateral exchanges of matter and energy in space, whereas this is imperative in hydrologic models of watersheds and wetlands (e.g. Costanza et al., 1990). Lateral exchanges, like soil erosion and water runoff, seldom are treated explicitly in point models focused on the temporal dynamics of soil C. The scope of soil C models varies immensely from single equation models to describe an individual process (e.g. first order decay of fresh plant litter) to whole ecosystem models which might include sub-models for plant growth; radiation and water transformation or exchange among air, plants and soil; energy and matter flows within the detrital food web; and organic matter decomposition and stabilization. The detail included in soil C models also varies immensely. For example, plant growth might be treated as an external input (i.e. regarded as outside of the modeled system's boundaries), or described as an empirical function of growing season rainfall, or modeled as the culmination of physiological processes down to the sub-cellular level.

Complex models of soil C dynamics

Most widely used soil C models are deterministic simulation models which tend to be relatively complex. Complexity arises from the following: integration of several rate equations describing the transfer of matter (and sometimes energy) from one pool (state variable) to another, modification of rates to adjust for sub-optimal environmental

conditions, several parameters to define key relationships (e.g. optimal rate coefficients), and input variables (e.g. temperature, precipitation, soil characteristics) to adjust for site-specific conditions. Model complexity also increases as greater detail on contributing processes is included, and this may require the time step of the model to be shortened. For example, some portions of the *ecosys* model developed by Grant (199?) operate on an hourly time step so that short-term variations (e.g. diurnal solar radiation) might be accommodated, but such large detailed models require considerable computing resources. Several models have been reviewed in the published literature, with earlier volumes tending to focus on soil N dynamics (e.g. Clark and Rosswall, 1981) and the more recent ones emphasizing C dynamics (e.g. Smith et al., 1998).

Perhaps one of the most widely used models of soil organic matter dynamics was developed at the Natural Resource Ecology Laboratory in Fort Collins USA, and dubbed Century, as it was originally intended to describe C cycling during 100 years. The model describes the flow of C and N from plant litter (partitioned between structural and metabolic components to reflect residue quality or decomposability) to soil organic matter which was fractionated among active, slow and passive pools to reflect decomposition and stabilization in the soil (Parton et al, 1982). Since its inception, Century has undergone continued development, and now uses a monthly time step (with a daily version being used by some developers), has the option to model P and S, and has been widely distributed and used in several biomes.

Another model which has persisted in the published literature even longer than Century is RothC, first described by Jenkinson and Rayner (1977). RothC regards plant residues as external inputs, but it is similar to Century in many other respects, including: use of a monthly time step, partitioning of plant materials into decomposable and resistant fractions, and fractionation of soil C among three pools (microbial biomass, humified organic matter and inert organic matter). Both Century and RothC have successfully described temporal changes in soil C for a wide range of environments. Of course being more comprehensive, Century is able to explicitly link C and N cycling as well as plant production, whereas RothC focuses solely on soil C dynamics.

Simple models of soil C dynamics

Simple models of soil C dynamics with analytical solutions have received considerably less attention since the 1960's when digital computers facilitated the development of simulation models based on numerical integration. Despite the ability of complex simulations to represent C cycling processes with ever increasing detail, scarce data are available to confirm that similar processes are important in real soils, and to independently estimate parameters and evaluate model output. The top-down approach attempts to model temporal changes in soil C with only as much detail, or the minimum number of parameters, required to capture the main features of soil C dynamics, such as the balance between plant C inputs and outputs of CO_2 from decomposition.

Soil C decay typically conforms to first order kinetics (i.e. decay rate is proportional to amount of decomposable substrate), and early models combined zero order additions of plant residues with first order decay to describe changes in soil C (Woodruff, 1949). The simple model used by Bremer et al. (1995) to describe C dynamics in soil light fraction (LF) was based on external inputs of plant C coupled with first order decay of C in the LF. More recently, an 'introductory carbon balance model' (ICBM) has been proposed by Andren and Kätterer (1997) as a simple alternative to more comprehensive simulation models. The model is solved analytically, regards inputs of plant C as an external variable, uses two pools of soil C (young and old) to describe decomposition and stabilization, and uses a single coefficient to adjusts decay and transfer rates for external climatic and soil factors.

How good are the models?

No single set of criteria has been accepted as the correct method of evaluating models. Various terms have been used to describe model evaluation: verification usually refers to mathematical correctness; confirmation refers to agreement between model and reality; validation usually means that the model has some capacity to represent or predict the behaviour of real systems, but also may imply that the model is legitimate, relevant or useful. Oreskes et al. (1994) assert that it is theoretically impossible to establish that a model accurately represents processes in real systems, and that the primary value of modeling is to help discover how systems might or might not function.

In theory models may only be evaluated by comparisons of output with data which

are completely independent of those used in model construction and parameterization, but this tends to be hampered by the lack of comprehensive data for a wide range of systems. Insight to model behaviour may be obtained from sensitivity analyses to assess which assumptions (including effects of uncertainty in parameter values, interactions among parameters, range of state variables, perturbations) are most crucial for model predictions. Sensitivity analyses, however, are difficult to manage for large models with several parameters, and typically indicate the linear dependence of model output on parameters even though most models are non-linear (i.e. the influence of a range of parameter values depends on one or more other parameters).

A search for a single 'valid' model for predicting temporal changes in soil C of Canadian agricultural land likely would be futile, because the predictive value of any model may not be certified and two or more different models may match most of the data equally well. The confidence in predictions tend to increase the more a model is tested and demonstrated to be consistent with reality, but this should not be confused for proof that the model assumptions are correct. Despite the implicit assumption that complex models perform better than simple ones, there is little evidence to suggest this is true. Contrary to the trend toward ever more detailed models built up from individual components and processes, Ågren and Bosatta (1990) argue for a higher level of abstraction in the development of general theories about ecosystem behaviour.

Summary and conclusions

Temporal changes in soil storage may be determined reliably for specific points in the landscape. The approach requires: a) representative sampling of the entire soil C pool, including coarse fragments of plant litter; b) interspersing of initial and subsequent samples to minimize the influence of spatial variations; c) accurate analyses of soil organic C concentrations; and d) comparisons based on an equivalent soil mass to adjust sampling depths for differences in soil bulk density. Despite providing useful information on soil organic C quality or decomposability, analyses of actively cycling fractions are difficult to standardize and the relationship between such fractions and total soil organic C may not be easily discerned. Thus such fractions might be used to assess soil organic C quality, but quantitative changes in C storage are more reliably based on total soil organic C.

The use of isotopic tracers at natural and artificially enriched levels to distinguish young soil C, recently derived from plant inputs, from older soil C is a powerful tool to investigate the production, decomposition, retention and stabilization of soil organic C. Isotopes are useful to assess the functional significance (e.g. whether turnover times differ as expected) of various fractions of soil organic C defined by physical, biological or chemical methods. Further studies of the decomposition and persistence of ^{13}C- or ^{14}C-enriched plant litter are required to assess the dynamics of plant residue C under contrasting management and environments in the field. Such studies should help to reduce uncertainties about the extent to which soil C storage might be manipulated through residue management.

Spatial variability in soil organic C within ecodistricts and individual landscapes likely necessitates the use of some model to estimate potential changes in soil C storage. The direct measurement of soil C change in all possible landscapes would be impractical and scientifically unrewarding. It seems doubtful that the error associated with aggregation of input variables justifies the use of complex simulation models to predict temporal changes in soil C. Perhaps using observed crop yields to estimate annual plant residue inputs might better reflect site-specific conditions and avoid the added uncertainty of simulated plant growth. Complex simulation models of plant production coupled with

soil C dynamics are valuable research tools to investigate interactions among ecosystem components and processes, but resulting estimates of temporal soil C changes may not be any more reliable than those from simpler models. Regardless of which model is selected to extrapolate from point measurements of soil C change to regional estimates, some assessment of probable errors in the estimates is essential

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